NASA-CR-202746

NASW-4574
7N-91-CR
004 931

Resurfacing history of Venus

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ABSTRACT

The stratigraphy, impact cratering record, and geoid of Venus indicate that a major change in surface geologic activity has occurred. Before ~800 Ma Venus was dominated by horizontal movement and tectonic deformation of the crust. A brief period of global volcanism subsequently filled in all low-lying areas and left much of the surface at a nearly uniform elevation. Current geologic activity is dominated by large volcanoes and limited horizontal movement of the crust. This geologic history may have resulted from mantle cooling causing the lithosphere to become positively buoyant globally.

INTRODUCTION

Radar images of Venus from the *Magellan* mission showed that the surface distribution of impact craters alone could not be distinguished from a spatially random population, that a small fraction (16%) of the craters are significantly modified by volcanism, tectonism, or erosion, and that the global crater production age is only ~500 m.y. (Phillips et al., 1992; Herrick and Phillips, 1994). These facts indicate a unique resurfacing history for Venus, and several hypotheses have been presented to explain them (e.g., Phillips et al. 1992; Schaber et al., 1992; Turcotte, 1993; Parmentier and Hess, 1992). All these hypotheses are devoted primarily to explaining limited aspects of the crater distribution and say little about the global stratigraphy and its relation to other data. Regional studies show a consistent global geologic stratigraphy, and there are recognizable relations of geologic features to the cratering record and the geoid.

GLOBAL GEOLOGIC RECORD

The landscape of Venus can be divided into three basic types of regions: tesserae, plains, and rift-volcano-coronae zones. A similar treatment of Earth would be to divide the surface into continents, oceanic plates, and hot spots. The tesserae are broad, pervasively deformed regions (Basilevsky et al., 1986). They are the only regions that indicate substantial horizontal tectonic deformation; some tesserae contain indenter zones and strike-slip faults with offsets >75 km (Pohn and Schaber, 1992). The tesserae are stratigraphically the oldest unit in any particular area, and globally tesserae underlie much of the Venusian plains (Solomon et al., 1992; Grimm and Phillips, 1992; Ivanov and Head, 1993).

Most of Venus's surface is made up of plains of volcanic origin located at a nearly uniform altitude (Slyuta and Nikolayeva, 1992). In many places there are no obvious volcanic sources for their emplacement. Almost all the plains contain at least modest tectonic deformation, some of which is coherent over vast areas (McGill, 1993). Much of the deformation is caused by proximity to structures such as rifts, volcanoes, and coronae (McGill, 1993); however, no plains deformation appears to be associated with tesserae.

Venus has several large shield volcanoes (flows >500 km in diameter) that are interconnected by a series of rifts that contain many coronae (Herrick and Phillips, 1992; Head et al., 1992). Coronae are 100–1000 km diameter volcano-tectonic structures with a ridged annulus (Basilevsky et al., 1986). Coronae are clustered about the mean planetary radius and appear to form passively as a result of rifting (Herrick and Phillips, 1992). The rift-volcano-coronae belts are stratigraphically the youngest features on the planet (e.g., Grimm and Phillips, 1992; Head et al., 1992), and their radar emissivity signatures indicate recent geologic activity (Robinson and Wood, 1993).

The stratigraphically young units on the surface exhibit evidence of only limited horizontal crustal movement. For example, the large rifts have not spread more than a few hundred kilometres. Likewise, subduction and large-scale underthrusting do not appear to be associated with the formation of the mountain belts around Lakshmi Planum (e.g., Hansen and Phillips, 1993) or the ridge belts in the northern plains.

IMPLICATIONS OF THE CRATERING RECORD

The small total number of impact craters on Venus limits their use for dating to broad scales. While the crater distribution alone cannot be distinguished from a spatially random population, and few of the craters have been modified since emplacement (Phillips et al., 1992), the craters were not emplaced on a geologically static surface. Phillips et al. (1992) found that regions of low crater density were statistically near embayed and tectonically deformed craters. Herrick and Phillips (1994) found that the crater distribution is nonrandom with elevation, denoted primarily by an excess of craters at the mean planetary radius and deficits at 1 km above and below it, and that the crater-deficient elevation ranges are associated with the rift-volcano-coronae zones.

Global geologic mapping, not yet done, is needed to denote the complete cratering record on different geologic units. However, the tesserae have already been extensively mapped, and their cratering record can be compared with the global mean (Table 1; tesserae map used is from Price and Suppe, 1993). The tesserae have more craters than would be expected from the global mean, but the excess is not statistically significant. The tesserae are at least as old as the mean age of Venus's surface. Ivanov and Basilevsky (1993) also located and measured the diameters of craters on the tesserae and compared their survey with a global crater database. They counted the same total number of craters on the tesserae as I did, but they counted eight more craters >16 km in diameter. They concluded that the tesserae are substantially older than the rest of the planet and that there must be many craters <16 km that are not observed. Without their data, explaining the difference between our results is speculative, but it may be that they used a different technique to measure crater diameter than I did or than that used in the survey (Schaber et al., 1992; Schaber and Chadwick, 1993) they compared their data to. The tesserae also have roughly three times the average percentage of embayed and tectonically deformed craters. Here, an embayed crater is one in which part of the floor or ejecta is covered by material from exterior sources. Typically, this means volcanic embayment, but in regions with extensive tectonic deformation (e.g., the tesserae), rubble from tectonic processes may cause the embayment.

Early modeling suggested that the seemingly random distribution of craters could be

TABLE 1. CRATERS IN TESSERAE COMPARED WITH THE TOTAL CRATER POPULATION

	Total	Craters in tesserae		Embayed (%)		Deformed (%)	
	observed	Expected	Observed	Total	Tesserae	Total	Tesserae
All	871	73	76	6.6	19.7	11.9	42.1
>16 km diameter	407	34	41	8.4	22.0	13.3	41.5
>32 km diameter	151	13	17	12.6	29.4	13.2	47.1

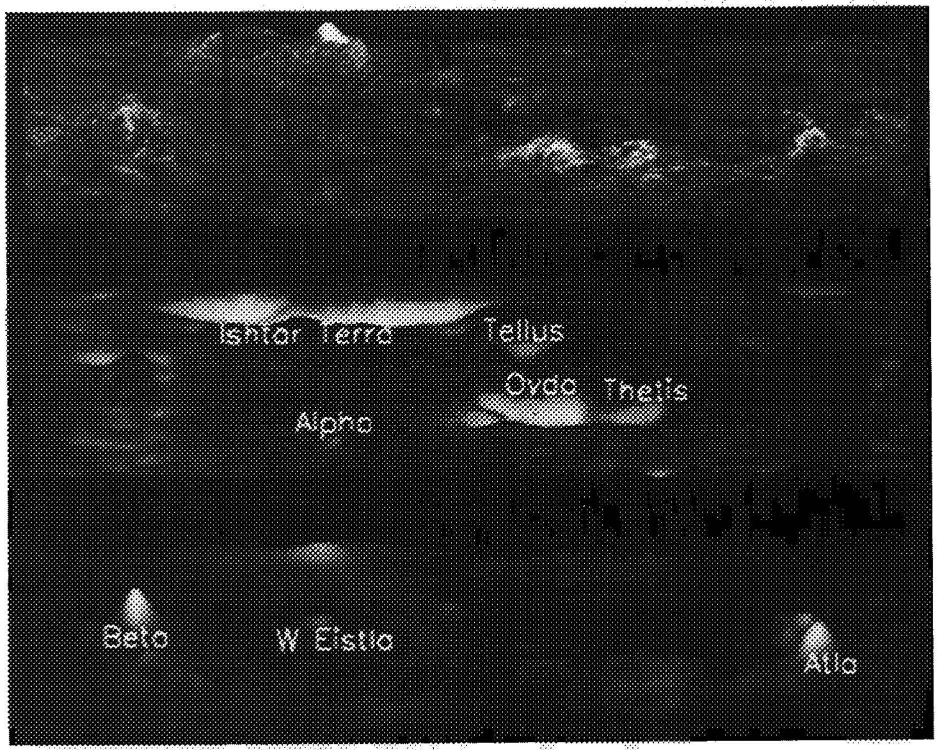

Figure 1. Three-layer perspective view showing surface of Venus and internal density structure at two different depths. Uppermost plane is global radar image draped over global topography. Lower two images represent inversion of spherical harmonic gravity and topography. Middle image shows density anomalies in lithosphere, and downwarped regions represent thickened crust. Bottom image shows density anomalies in upper mantle, and upwarped regions represent mantle upwellings. Major tesserae regions (e.g., Alpha, Ovda, Thetis, Tellus) are all regions of thickened crust, whereas major volcanic regions (e.g., Atla, Beta, W. Eistla) are mantle upwellings. Image covers area from long 120°W to 240°E and lat 80°N to 50°S.

the result of infrequent resurfacing over very large areas or frequent resurfacing over small areas (Phillips et al., 1992). Other modeling suggested that the latter would have produced too many embayed craters and that at most 10% of the planet has been resurfaced since a planet-wide event (Schaber et al., 1992). Although these models have value, they are two-dimensional in nature and valid only if resurfacing is spatially random, periodic with time, and in single-sized circular patches. For example, a three-dimensional model with a size distribution of volcanic features similar to that of the present can reproduce the observations with catastrophic resurfacing at 550 Ma followed by additional resurfacing that produced the equivalent of a 440-m-thick global layer (Bullock et al., 1993). Assuming an initially craterless surface, the observed deficits with elevation require that ~30% of the planet was resurfaced with enough material to bury preexisting impacts, if resurfacing has occurred at a constant rate.

IMPLICATIONS OF THE GLOBAL GEOID

Direct information about the interior of Venus comes from spacecraft gravity data, commonly expressed as a spherical harmonic expansion of the gravitational potential. The best available expansion is to degree and order 60 and includes data from *Magellan's* fourth cycle (Konopliv et al., 1993). The geometry of the orbits of *Magellan* and *Pioneer Venus* cause the data to be of decreasing reliability as one goes north or south from about lat 10°N. At the wavelengths of the available data (>600 km), surface features currently influenced by the mantle convection pattern should have a sizable geoid signature associated with them. For example, Earth's long-wavelength geoid is dominated by the signatures of subduction zones and hot spots (Richards et al., 1988).

Herrick and Phillips (1992) inverted low degree and order spherical harmonic expansions of topography and the geoid (from *Pioneer Venus* data) to separate major mass anomalies within the lithosphere from those within the mantle; they then compared those results with observed surface features. The mass anomalies were represented by surface density anomalies on two spherical shells placed at appropriate depths within the planet. The technique used by Herrick and Phillips (1992) basically separated topographic features with small (upper shell) and large (lower shell) geoid signatures. Here, I use their model (assuming an isoviscous mantle) with the more recent, higher resolution expansion of the potential to perform the inversion to degree and order 30 (Fig. 1). The top image in Figure 1 is a global radar mosaic draped over the topography, the middle image is a shell at 30 km depth, and the lower image is a shell at 200 km depth. Negative surface density anomalies on the upper shell, which can be interpreted as regions of thickened crust, are represented by depressions. Negative surface density anomalies on the lower shell, which can be interpreted as regions of upwelling mantle material (plumes), are upward protrusions.

Several distinct correlations exist among surface morphologic features and the two shells. Perhaps most striking is that every major block of tesserae shows up as thickened crust in the upper shell. In contrast, there is no obvious correlation between the lower shell and tesserae, though there are a few tesserae regions, particularly in Ishtar Terra, that appear to have signatures in the lower shell. Possible interpretations are that tesserae regions with signatures in the lower shell are actively forming (Bindschadler et al., 1992), or they are fortuitously located over an unrelated mantle density anomaly, or the model's inherent assumption of a single global lithospheric thickness is incorrect and causes the apparent signatures. Also, the inversion assumes that the absolute values of the gravity data are correct, which may not be the case at the latitude of Ishtar Terra (Konopliv et al., 1993). The lower shell correlates quite strongly with the locations of large volcanic structures on Venus. At first glance this is not surprising, as Earth's geoid is well correlated with the hot-spot distribution (Richards et al., 1988). However, the correlation on Earth exists only if active volcanic structures are compared to the geoid. The high correlation on Venus implies that most of the large volcanic centers have been active in the geologically recent past.

PROPOSED GEOLOGIC HISTORY

Analysis of the stratigraphy, the impact cratering record, and the geoid indicates that a major change in the nature of Venus's surface geologic activity has occurred within the past 1 b.y. (all ages are referenced to a mean crater retention age of 550 Ma; error bars on this age are a factor of 2). The tesserae are remnants of a past era that had substantial horizontal movement and tectonic deformation of the crust, followed closely by rapid emplacement of volcanic plains that filled in the low-lying areas of the planet. The current era is dominated by hot-spot tectonics (Phillips and Malin, 1983), where upwelling mantle plumes and their associated lithospheric stresses produce large volcanoes and limited horizontal movement of the lithosphere.

Many observations fit nicely with the idea that tesserae are remnants of a previous geologic era. Because tesserae are not currently forming, they have no substantial geoid signature and are at the bottom of the stratigraphic column. Burial of the low-lying regions of tesserae explains why they exhibit extensive deformation but few plate-boundary structures. If tesserae were highly active and then quickly "froze," existing craters at shut-off time would be deformed and embayed, while later craters would be emplaced but not modified. If cessation of tesserae deformation was followed shortly by rapid emplacement of the plains, then the tesserae would have slightly more craters than the plains and substantially more modified craters. The observed relative number of craters and the embayment percentages (Table 1) are consistent with the tesserae having a mean retention age of 100 m.y. (where all existing craters are embayed) before shut-off time, followed by plains emplacement within the next 50 m.y. However, the high percentage of tectonically deformed craters on the tesserae is inconsistent with this scenario. Some form of modest tectonic deformation, such as gravitational relaxation, may have occurred since plains emplacement. Ivanov and Basilevsky (1993) found a smaller percentage of craters on the tesserae (17% of craters >32 km) with significant tectonic deformation.

The similarity in age to the tesserae, the low number of embayed craters, and the uniform crater distribution suggest that plains emplacement was both rapid and globally nearly simultaneous. The uniform elevation and lack of large, visible volcanic sources over much of the plains indicate a flood-basalt emplacement mechanism. Emplacement globally as a single unit also helps to explain why subsequent surface deformation is spatially coherent over large areas. Because the tesserae have not deformed the plains, at least the final stages of plains emplacement must postdate cessation of significant tectonic activity in the tesserae. The cratering record suggests that plains emplacement was completed 600–700 Ma.

The stratigraphy dictates that the rift-volcano-coronae zones postdate most plains emplacement. The cratering record, the geoid, and the emissivity signatures suggest that these regions are currently geologically active. The high correlation of the geoid to large volcanic structures suggests that there has been minimal reorientation of the mantle convection pattern since the large volcanic structures began forming. The stratigraphically young rifts, ridge belts, and mountain ranges are indicative of a lithosphere that is too buoyant for subduction but that has limited horizontal movement in response to shear tractions from mantle convection (Phillips et al., 1991). The cratering record indicates that hot spots have removed 15%–30% of the craters emplaced since the plains were formed.

POSSIBLE PHYSICAL MECHANISM

The geologic history can be summarized as a shift from a highly mobile crust to a largely immobile crust, with a brief period of volcanic flooding in the interim. If it is assumed that large-scale crustal movements are controlled by the lithosphere as the boundary layer of mantle convection, then the geologic history suggests a change in the boundary condition from free slip to no slip. With this principle, I propose that prior to ~800 Ma Venusian plate tectonics existed, but ceased when mantle cooling caused the lithosphere to become positively buoyant. Here the term "plate tectonics" does not imply a set of rigid plates as it does for Earth, but it does imply crustal recycling through subduction and spreading. After the cessation of plate tectonics, the no-slip boundary condition caused the mantle to heat up, raising the temperature of the lithosphere's base above the solidus. With mantle material being continuously brought to the base of the lithosphere, a brief period of global volcanism ensued, ending when the thickness of the crust and the buoyant residuum reached the lithosphere's thickness. After global flooding, volcanism occurred primarily over mantle plumes. The proposed model requires that (1) some areas of the planet previously had a negatively buoyant lithosphere, (2) when the lithosphere became positively buoyant planetwide, the mantle temperature quickly raised the base of the lithosphere above the solidus temperature, and (3) volcanic flooding lasted for a brief time.

Using a one-dimensional parameterized convection model, Phillips and Malin (1983) showed that a lithosphere in steady state with the Venusian mantle is likely to be positively buoyant and concluded that subduction did not occur in the past. While their work demonstrates that the Venusian lithosphere is generally more buoyant than Earth's, their model has certain limitations that disallow ruling out past subduction. Subduction requires only that old, cooled, Venusian lithosphere (as opposed to "average" lithosphere) be negatively buoyant. Also, their model requires that the lithosphere's base be defined by a fixed temperature (1500 °C) rather than a fixed fraction of the mantle convecting temperature.

On Earth, only oceanic lithosphere is negatively buoyant, and it becomes so by cooling while moving away from a spreading center. Although more sophisticated techniques exist to estimate the density of the lithospheric column, for the discussion here it is adequate to use a simple analytical approximation of the mantle-lithosphere density contrast (Oxburgh and Parmentier, 1977):

$$\Delta\rho = \frac{y_c}{y_l}[-\rho_c + \rho_m] + \frac{y_r}{y_l}[-\rho_r + \rho_m] - \frac{1}{y_l}\alpha\rho_m\,(T_m - T_s)\frac{2(\kappa t)^{1/2}}{\sqrt{\pi}},$$

where y_c, y_r, and y_l are crustal, residuum, and lithospheric thickness, ρ_c, ρ_r, and ρ_m are unheated crust, residuum, and mantle density, α is the coefficient of thermal expansion, $(T_m - T_s)$ is the mantle-surface temperature contrast, κ is thermal diffusivity, and t is the age of the lithosphere. If the base of the lithosphere is taken as the point where the temperature T is such that $(T - T_s)/(T_m - T_s) = 0.9$, then $y_l = 2.32(\kappa t)^{1/2}$ (Turcotte and Schubert, 1982, p. 160). Here, I assume that $\rho_c = 3000$ kg/m^3, $\rho_r = 3290$ kg/m^3, $\rho_m = 3300$ kg/m^3, $T_s = 735$ K, $y_r = 2.5\,y_c$, and $\kappa = 10^{-6}$ m^2/s. Figure 2 plots the boundary between positive and negative buoyancy as a function of y_c and T_m for a lithosphere at 20, 40, 60, and 80 Ma. For a free-slip boundary, parameterized convection models give a possible T_m range of ~1450–1800 K (50–200 K higher than on Earth) and estimate that the mantle has cooled a few tens of degrees over the past 2 b.y. (e.g., Phillips and Malin, 1983; Turcotte, 1993). A hotter mantle temperature on

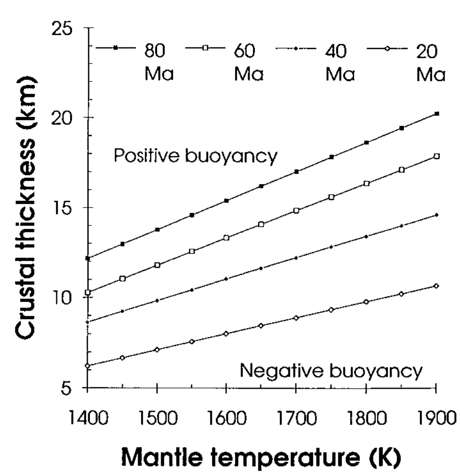

Figure 2. Boundary between positive and negative buoyancy for lithosphere of different ages as function of crustal thickness and mantle temperature. Lithosphere is negatively buoyant and, thus, capable of subducting if it has higher mean density than underlying mantle.

Venus than on Earth makes it likely that spreading processes would produce more crust by a factor of 2 or 3 (Sotin et al., 1989). As plotted in Figure 2, the range in acceptable values for T_m and y_c indicate that the buoyancy of the Venusian lithosphere could have changed from negative to positive within the past 1 b.y.

If the boundary condition in a parameterized convection model is changed from free slip to no slip, then a rapid and substantial rise in the mantle convecting temperature occurs. Using a parameterized convection code based on Turcotte et al. (1979), I estimate that the mantle convecting temperature would rise about 150 K, and that 100 K of that would occur in 1 b.y. and 30 K in the first 150 m.y. after the change in boundary condition. Oceanic lithosphere on Earth does not generally appear to cool to a thickness greater than about 100 km (Johnson and Carlson, 1992); i.e., the globally averaged temperature under oceanic lithosphere is T_m. Once subduction ceases, the only way for large-scale volcanism to occur is if the base of the lithosphere, which is now a conducting lid, is heated above the solidus. Thus, if T_m rises above the solidus for 100 km depth, widespread volcanism will occur. With convective velocities on the order of 1 cm/yr, within a few tens of millions of years the lithosphere will be entirely crust plus mantle residuum. After this episode, volcanism will occur primarily at locations of mantle upwelling. Given that the solidus for mantle material at 100 km depth is only about 1600 K (e.g., Parmentier and Hess, 1992), it is conceivable that a 30 K rise in mantle temperature could cause widespread melting.

DISCUSSION

When taken as a whole, the stratigraphic, cratering, and geoid data clearly indicate that geologic activity on Venus underwent a fundamental change. The crust was globally highly mobile and heavily deforming until ~800 Ma. Cessation of deformation was closely followed by global volcanic flooding. The present era shows geologic activity concentrated around large volcanic centers interconnected by rifts with limited horizontal crustal movement. I propose that this geologic history results from the cessation of plate tectonics and a subsequent shift to hotspot tectonics; mantle cooling made the lithosphere positively buoyant. The proposed model predicts that no tesserae are currently forming, that all plains are nearly the same age, and that horizontal movement of the surface is currently limited to a few hundred kilometres, predictions that can be tested by regional geological and geophysical analyses. The rapidly improving sophistication of convection modeling will provide insight into what happens when a mantle cools to the point where the lithospheric column (crust and mantle) is no longer negatively buoyant.

ACKNOWLEDGMENTS

Supported by NASA grant NASW-4574 to the Lunar and Planetary Institute. I thank R. E. Grimm, W. Kiefer, and R. J. Phillips for discussions; M. Price for supplying a global tesserae map; K. Herrick for technical editing; and T. Watters for his review. LPI contribution 833.

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Manuscript received January 3, 1994
Revised manuscript received May 19, 1994
Manuscript accepted May 25, 1994

Printed in U.S.A.